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RECENT TRANSFORMATIONS IN THE HIGH-ARCTIC GLACIER LANDSYSTEM, RAGNARBREEN, SVALBARD

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ABSTRACT. Ragnarbreen is a small glacier located in the central part of Spitsbergen Island (Svalbard archipelago) and fed by the larger ice mass of Mittag-Lefflerbreen. Glacier recession and landform development in the foreland of Ragnarbreen are quantified using time-series orthophotos and digital elevation models, which were generated based on aerial photographs from 1961 (black and white frame camera), 1990 (false infrared frame camera) and 2009 (colour digital camera), obtained from the Norsk Polar Institute.

Receding from its maximum Little Ice Age extent, attained in the period 1900/1920–2013, the glacier margin retreated by 1658 m while the extent of the area of ice decreased by 26%. The glacier snout lost 135 million m^3 of ice during the period 1961–2009, whereas landform changes (mainly due to dead-ice melting and debris flow activity) were more than twenty five times lower, with the less than 5 million m^3 of sediment and dead ice volume loss. In terms of landscape alteration between 1961 and 2009, the most important was the creation of a terminoglacial lake, which acted as a sedimentary trap and at the same time probably accelerated glacier retreat. The second most active component was the lateral moraines whose transformations were divided into four phases, with various magnitudes of debris flow and backwasting activity that changed with time. The end moraine complex was the most stable component, affected mainly by dead-ice downwasting and to a lesser extent by sporadic debris flows.

Key words: ice-cored moraine, Spitsbergen, glacial geomorphology

Introduction

The extensive recession of Svalbard's glaciers provides a unique opportunity for the study of landscape alteration in response to changes in climatic conditions. Quantification of process-form relationships is especially important in the context of predicting and modelling of future environmental changes. Most studies related to the quantitative analysis of environmental transformations on Svalbard have concentrated on changes in glacier geometry, dynamics and mass balance (Hagen and Liestøl 1990; Hagen *et al.* 1993; Jania and Hagen 1996; Melvold and

Hagen 1998; Bamber *et al.* 2005; Nuth *et al.* 2007; Rachlewicz *et al.* 2007; Sund *et al.* 2009; Moholdt *et al.* 2010a; 2010b; Nuth *et al.* 2010; Kristensen and Benn 2012; Murray *et al.* 2012; Małecki *et al.* 2013). Much less attention has been paid to alterations in glacial landforms including changes due to dead-ice melting (e.g. Bennett *et al.* 2000; Etzelmüller 2000a, 2000b; Lønne and Lyså 2005; Lukas *et al.* 2005; Schomacker and Kjær 2008; Irvine-Fynn *et al.* 2011).

The aim of the study is to quantify decadal landscape changes in the area undergoing recent deglaciation. This study presents quantitative data on the transformation of Ragnarbreen, Svalbard, based on multi-temporal digital elevation models (DEMs) constructed from aerial photographs. Furthermore, dominant geomorphic processes within the glacier's foreland are identified and related to the overall landscape changes. A model of glacier and landform transformations is also presented.

Regional settings

The study area is located in the Svalbard archipelago in a high-Arctic setting. Almost 60% of the archipelago is covered by glaciers, containing ice caps, ice fields, cirque and valley glaciers (Nuth *et al.* 2013). The Little Ice Age terminated on Svalbard around 1920 and significant rise of temperature occurred during 1920s and 1930s (Nordli and Kohler 2003; Nordli *et al.* 2014). Thus, many studies assumed that most of Svalbard's glaciers started to retreat from their Neoglacial maximum extent around the 1920s (Lefauconnier and Hagen 1990; Hagen *et al.* 1993). However, field observations provided by reports from the early 20th century arctic expeditions - which documented that during 1910s and 1920s some glaciers had already retreated from moraines marking its maximum extent (cf. Lamplugh 1911; Slater 1925) - suggest that some of the glaciers started to retreat earlier, around the 1900s.

Ragnarbreen is located north of Petuniabukta at the head of Billefjorden on the island of Spitsbergen (Fig. 1). The climate of the study area is more continental than the western coast of the island, with a smaller amount of precipitation and slightly higher variations in temperature amplitude (Rachlewicz and Styszyńska 2007; Rachlewicz 2009). As a consequence, the extent of the ice cover around Petuniabukta is also smaller.

Previous work on glaciers and glacial landforms in the study area concerns both geomorphology (e.g. Kłysz 1985; Gonera and Kasprzak 1989; Karczewski 1989; Stankowski *et al.* 1989; Karczewski *et al.* 1990; Karczewski and Kłysz 1994; Gibas *et al.* 2005; Rachlewicz and Szczuciński 2008; Rachlewicz 2009; Hanáček *et al.* 2011; Evans *et al.* 2012) and glaciology (e.g. Rachlewicz *et al.* 2007; Rachlewicz 2009; Małecki 2013; Małecki *et al.* 2013).

Ragnarbreen is a small glacier flowing from the larger ice mass, Mittag-Lefflerbreen. The four main landscape elements of Ragnarbreen are: glacier snout (clean glacier surface), terminoglacial (partly supraglacial) lake, lateral moraines, and an end moraine complex. Processes and sediments of the last component have been described previously (Ewertowski *et al.* 2012), so that in this study the main focus is on the transformation of the glacier snout, lake and lateral moraines.

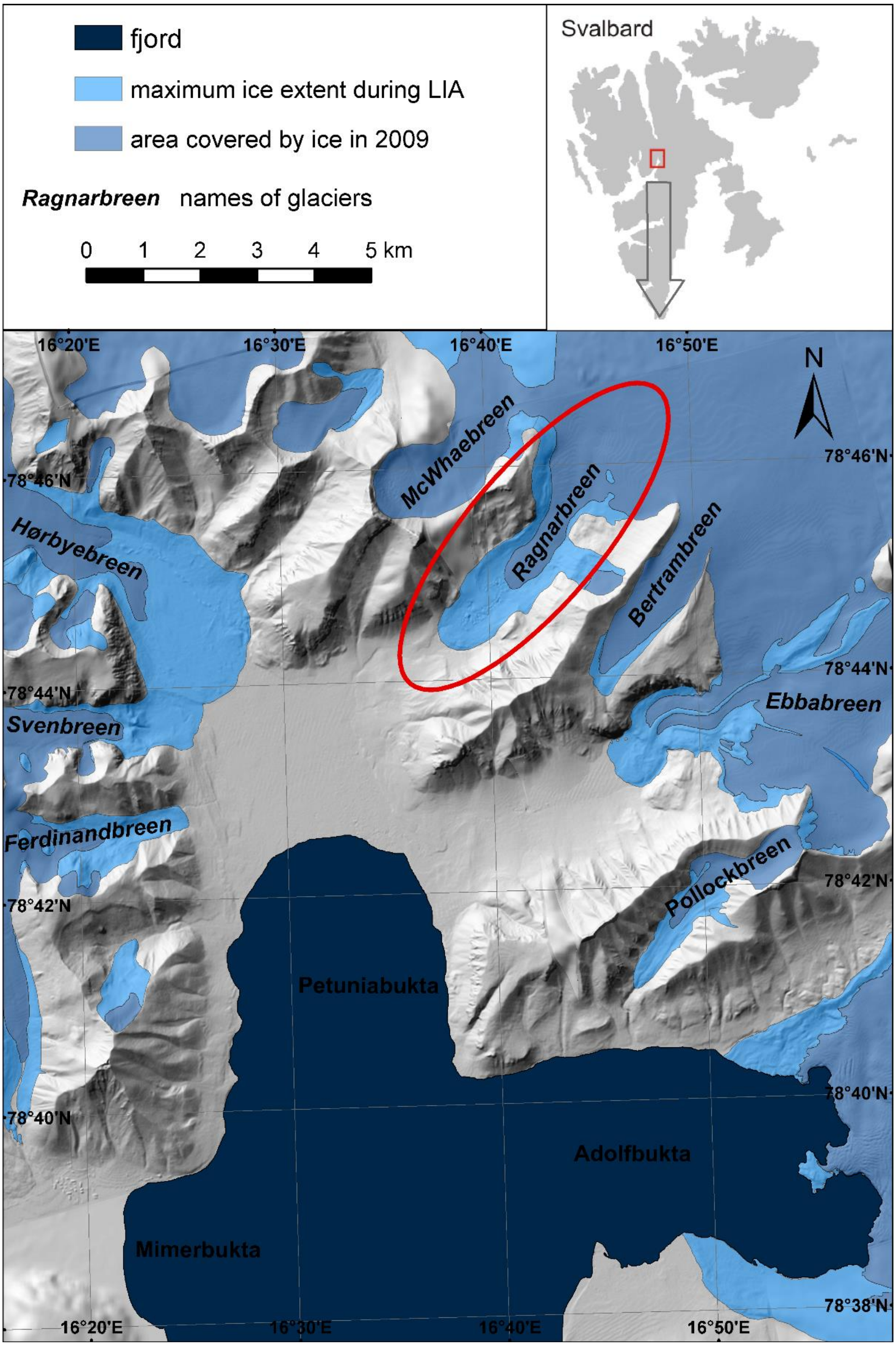

Figure 1. Location of the Ragnarbreen (circled)

Methods

Landscape changes were quantified based on multi-temporal aerial photographs and digital elevation models (DEMs) created from aerial stereo pairs. Three sets of aerial photographs were obtained, courtesy of the Norsk Polar Institute (NPI), and processed using photogrammetric software for the generation of orthophotos and DEMs:

1. Black and White frame camera photographs from 1961: S61-3102 (F72), S61-3103 (F-73) and S61-3104 (F74). The ground resolution of these photographs was 0.55 m.
2. False infrared colour frame camera photographs from 1990: S90 1904, S90 1905, S90 1906. The ground resolution of these photographs was 1 m.
3. Colour digital camera photographs from 2009: S2009 13833_0481, 13833_0482, 13833_0483, 13833_0486, 13833_0487. The ground resolution of these photographs was 0.4 m. The narrow strip of the NW part of the lateral moraine was only covered by one of the photos; hence, for this area, DEM construction was not possible.

The projection used was UTM zone 33N. Coordinates of control points were collected around the glacier's foreland with the use of dGPS during summer of 2013. In total, 42 points were surveyed, 25 were used as ground control points (GCP) to establish absolute orientation for orthorectification of aerial photographs and DEM creation, whereas a further 17 were used as independent checkpoints to assess the quality of the processed data (i.e. they were not used for orthophotos and DEMs creation, but to evaluate the horizontal and vertical errors of the final products). The root mean square error (RMSE) in 3D space compared to ground surveyed checkpoints was as follow: 1.16 m for 1961, 1.48 m for 1990, and 0.95 m for 2009, with horizontal errors slightly larger than vertical ones (Table 1). Moreover, systematic and random vertical errors were assessed by comparison of 1961 and 1990 DEMs with the 2009 DEM at points where elevation was expected to be stable (i.e. non-glaciated, non-ice-cored, non-dynamic rock areas) (Table 1). These errors are higher than these calculated from ground-surveyed checkpoints with RMSE 2.60 for 1961 and 3.75 for 1990. Error distribution was close to normal suggesting random errors rather than systematic. Larger errors for 1990 products resulted from coarser resolution of 1990 aerial photographs. Calculation of volume changes in landforms and glacier snout was done with the use of DEMs of Differences (DoDs) (cf. Wheaton *et al.* 2010). i.e. generated DEMs were subtracted from each other, enabling a spatial picture of the loss or deposition of material to be obtained in each cell of the model from one period to another. Then the values of elevation changes were multiplied by cell size, and thus the volume changes were calculated. All elevations provided in the text are ellipsoidal. Lake depth was surveyed in 2005 from rubber boat using echo sounding with GPS and crosschecked by manual depth soundings (using rope with weight).

Geomorphological features were identified in the orthophotos aided by visualization of the DEMs (relief shading with different light directions) and its derivatives (slope gradient, aspect and curvatures) Additional field geomorphological mapping and ground checking were carried out during fieldwork in 2005, 2007, 2012 and 2013. Geomorphological maps were compiled and digitally drawn in a GIS environment. They will be published in the near future together with maps concerning neighbourhood glaciers.

Table 1. DEMs specification and errors based on independent ground survey checkpoints and internal checkpoints (2009 DEM)

Year	DEM cell size (m)	Accuracy compared to ground surveyed checkpoints (n = 17)			Vertical accuracy and precision of DEMs compared to 2009 checkpoints		
		Vertical RMSE	Horizontal RMSE	Total (3D) RMSE	Mean error	Standard deviation of error	RMSE
		(m)					
1961	1.0	0.65	0.94	1.16	1.25	2.28	2.60
1990	1.0	0.80	1.27	1.48	1.19	3.55	3.75
2009	0.4	0.63	0.76	0.95	-	-	-

Results

Spatial and temporal trends in evolution of the glacier snout and terminoglacial lake

*Glacier extent changes during recession from its LIA maximum position: 1900(1920)–2013***.** The maximum LIA extent of the glacier was calculated using lateral and end moraine complexes, it was assumed that the outer edges of these moraines marked the approximate position of the ice margin. As there is no direct evidence about the timing of recession, it was assumed that the glacier started to retreat from this position around either 1900 or 1920 (similarly to the other glaciers on Svalbard) and thus these two dates were used for further calculations. Glacier extents from 1961, 1990 and 2009 were distinguished based on aerial photographs. In the upper part of the glacier, the border between Ragnarbreen and neighbouring glaciers was delineated based on elevation changes represented by ice surface contours. The total decrease in the glacier's extent over the period 1900/1920 and 2009 was 3.33 km^2 or 26% of the maximum extent (Table 2, Fig. 2a). It has to be stressed that the delineated area of ice refers to the glacier surface without debris cover, because there is still a lot of ice buried under the debris cover within the moraines.

Table 2. Changes in glacier extent, Ragnarbreen, Svalbard. Note that values of area changes are provided jointly for the whole glacial (i.e. main glacier body plus small niche glacier)

Year	Area of ice cover (km^2)	Percentage area of ice cover remaining (%)
LIA maximum (1900/1920)	13.05	100
1961	11.78	90
1990	10.62	81
2009	9.72	74

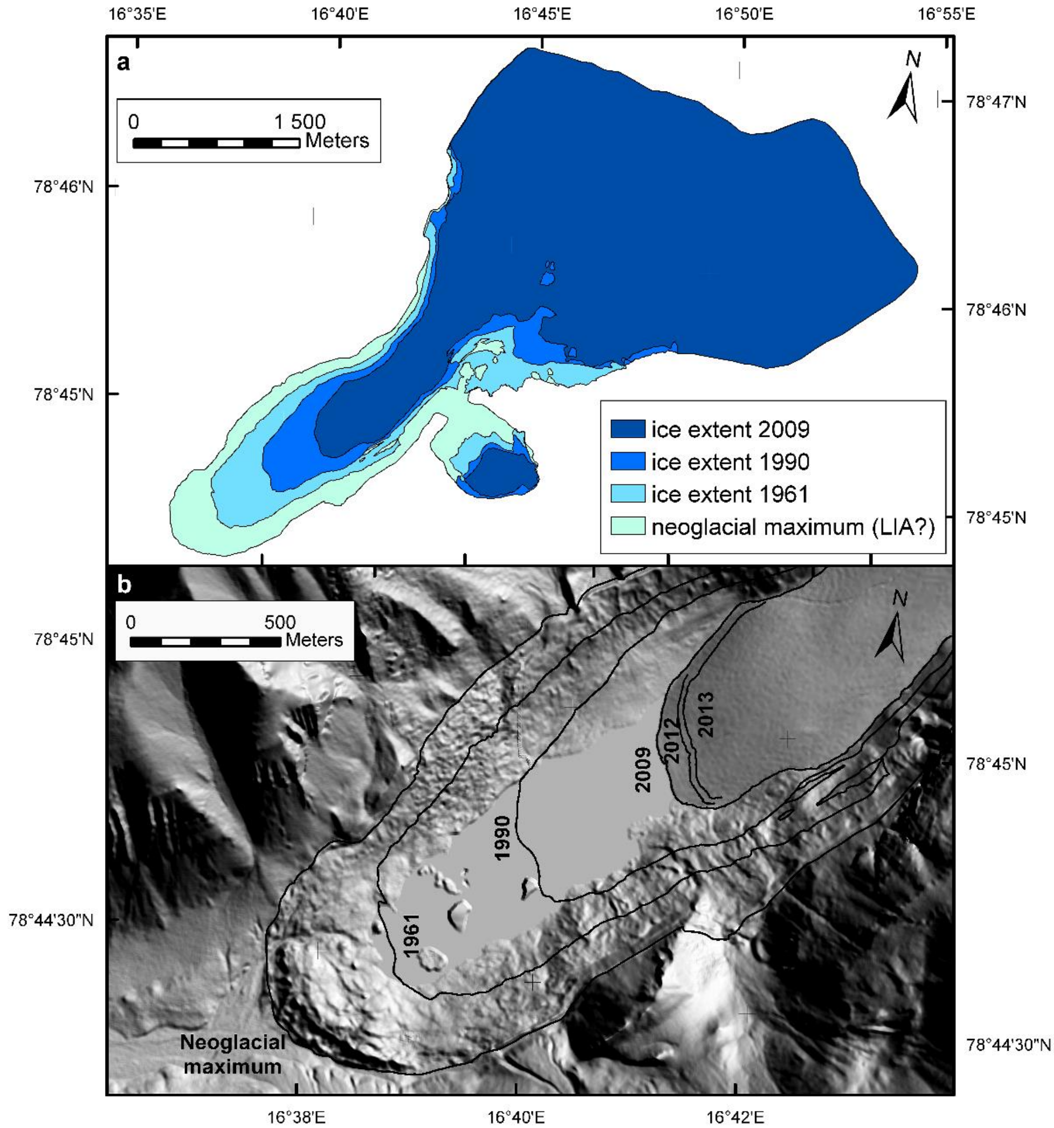

Figure 2. (a) Changes in the extent of the ice, and (b) changes in glacier margin position (Ragnarbreen, Svalbard). Note that extent changes are provided for the whole glacial basin jointly.

The recession of the glacier margin is especially visible for the glacier snout (Fig. 2b). The snout retreat from its maximum extent to the 2013 ice margin position was 1658 m and the mean annual retreat rate was 14.7 m a^{-1} (1900) or 17.8 m a^{-1} (1920) depending on which date will be taken as the beginning of glacier recession. However, recently the retreat rate has increased (Table 3). The values of the snout recession were measured along one line following the valley and glacier's main axis. Thus, they are slightly different from results shown in Rachlewicz et al. (2007), where the maximum distances between the glacier's extent for each period were provided.

Table 3. Recession of glacier margin from its Neoglacial maximum extent: 1900/1920 -2013, Ragnarbreen, Svalbard. Note: The values of the snout recession were measured along one line following the valley and glacier's main axis.

Time period	Glacier margin recession	
	Total (m)	annual mean retreat rate (m a^{-1})
Neoglacial maximum extent (1900/1920)–1961	453	7.4/11.0
1961–1990	557	19.2
1990–2009	534	28.1
2009–2012	92	30.7
2012–2013	22	22.0
Neoglacial maximum extent (1900/1920)–2013	1658	14.7/17.8

Volume and elevation changes within glacier's foreland and snout (1961–2009). The glacier surface in profile parallel to ice flow was significantly lowered during the period 1961–2009 by up to 90 m (Fig. 3b). An additional ~17 m can be added for the area where currently there is the deepest part of the terminoglacial lake. In 1961, the glacier surface showed a characteristic steep ice front (mean slope 11%), followed by a flatter section (over the contemporary lake) with a mean slope of 5%. In 1990 and 2009, the glacier snout was characterized by a mean slope of 8% and 10% respectively, with no flat section. The glacier's thickness in several profiles, transverse to ice flow, was also greatly lowered. In the transverse profiles (Fig. 3c), ice surfaces showed slight asymmetry, with the NE side being lower than the SW side.

Detailed DEMs of Differences were used to calculate the volume and spatial distributions of changes during the periods 1961–1990, 1990–2009, and 1961–2009 (Fig. 4). The total volume lost from the glacier's foreland and snout (i.e. decrease in the volume of active ice plus decrease in volume of debris covered ice) over the period 1961–2009 was ~140 million m^3. The majority of this loss (> 135 million m^3) was related to active ice melting. Mean annual changes in the snout area and volume vary over the study period (Table 4); however, a general tendency towards increased volume loss was visible. Spatial variability was also apparent: the NW part of the glacier was most reduced, which is reflected in the pattern of elevation changes (Fig. 4). The maximum value of elevation changes during the period 1961–1990 for the NW part of the glacier snout was –2.16 m a^{-1}, whereas for the period 1990–2009 it was –2.81 m a^{-1}. Spatial variability in elevation changes resulted in an asymmetric transverse profile (Fig. 3).

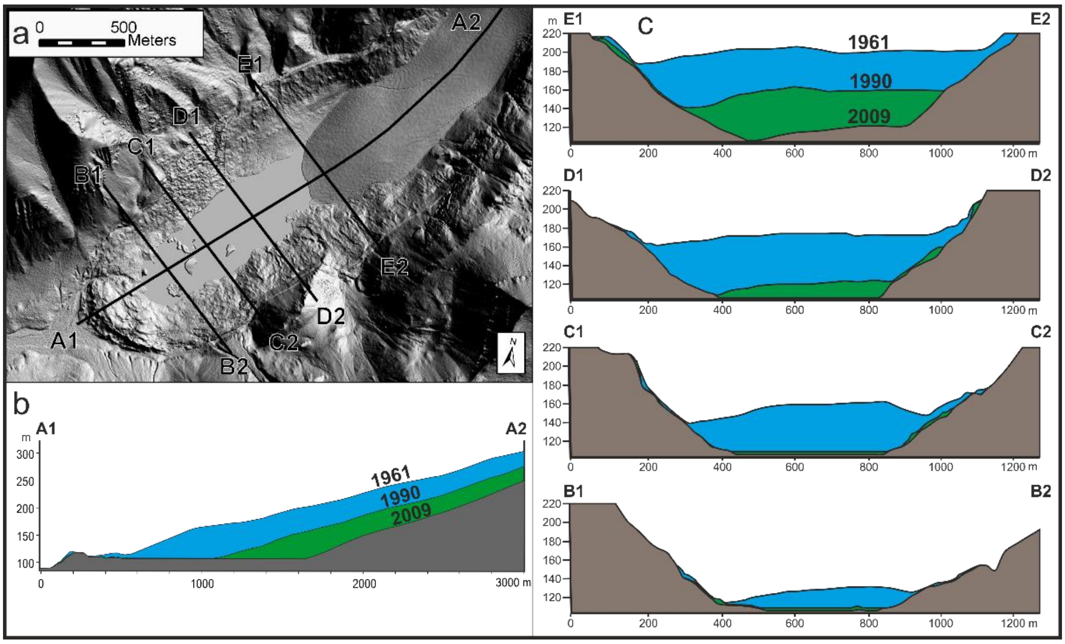

Figure 3. Surface changes of the glacier's foreland and snout, Ragnarbreen, Svalbard: (a) profiles' locations; (b) profile A1–A2, parallel to the ice flow direction; (c) profiles transverse to the ice flow direction.

Table 4. A comparison of volume changes over time. Landform (dead-ice + debris) volume was calculated for the foreland exposed after 1961 only.

End moraine complex						
	elevation difference					
Time period	maximum	minimum	mean	annual mean	volume change	area
	m	m	m	m a^{-1}	m^3	m^2
1961–1990	3.2	−10.2	−1.6	−0.055	−396 454	261 899
1990–2009	2.1	−11.8	−0.1	−0.003	−13 468	261 899
1961–2009	4.2	−11.4	−1.6	−0.033	−409 921	261 899
Lateral moraines exposed after beginning of the recession and before 1961						
1961–1990	4.3	−14.7	−4.2	−0.15	−1 834 730	432 372
1990–2009	4.9	−9.3	−1.6	−0.08	−671 512	432 372
1961–2009	5.7	−16.5	−5.8	−0.12	−2 507 445	432 372
Lateral moraines exposed between 1961 and 1990						
1990–2009	3.4	−28.8	−3.1	−0.16	−1 837 816	587 694
Glacier snout						
1961–1990	0	−62.7	−32.9	−1.14	−85 020 046	2 581 266
1990–2009	3.8	−53.4	−23.4	−1.23	−46 713 675	1 993 305
1961–2009	1.5	−99.2	−50.4	−1.05	−135 299 426	2 685 755

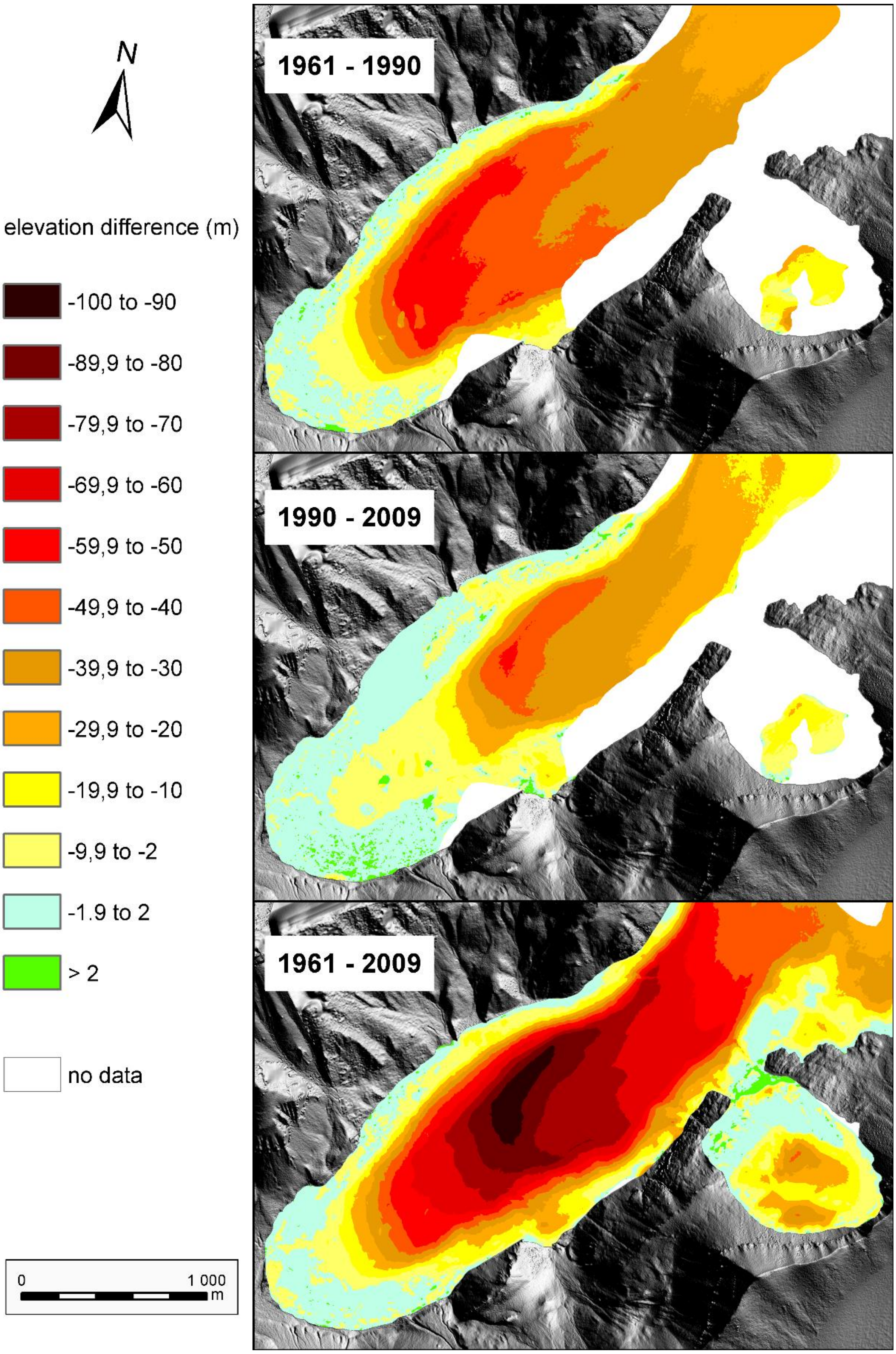

Figure 4. DEMs of Differences (DODs) showing the spatial distribution of surface transformation of the glacier snout, Ragnarbreen, Svalbard. Areas assumed to be stable or whose transformation was within the range of DEM errors are marked in light blue

Evolution of terminoglacial lake and drainage system (1961–2013). In 1961, the main supraglacial drainage was flowing on the SE part of the glacier. On the north side, only smaller supraglacial streams were observed. A subglacial outflow was located immediately behind the end moraine complex with the stream running through the complex. There was no big lake in front of the glacier margin. A terminoglacial lake started to develop between 1980 and 1984 according to personal communication from members of the 1979 and 1984 expeditions (pers.com. L. Kasprzak, P. Kłysz, 2004). This distal part of the lake (formed between 1961 and 1990) was shallow (up to 2 m deep) and contained quite a number of islands. The islands' maximum height above the lake level was 3 m, but usually did not exceed 1 m above the lake level. Two small ponds have developed inside two of the islands.

In 1990, the main supraglacial stream was located in the central and NW part of the glacier. The terminoglacial lake was 780 m long and more than 400 m wide (Table 5). The lake has subsequently grown, reaching a length of 1400 m in 2013. The main drainage from the lake was in a similar position during the study period; however, the lake level and in consequence the extent changed slightly over the years (Table 5; Fig. 5).

Table 5. Morphometry of terminoglacial lake, Ragnarbreen, Svalbard

Year	Maximum length (m)	Maximum width (m)	Maximum depth (m)	Area (m^2)
1990	778	444	–	221 141
2005	1080	450	17	390 000
2009	1320	472	–	431 336
2012	1383	472	–	461 165
2013	1405	472	–	470 034

Since 1990, the proximal part of the lake has developed. The proximal part was also deeper than its distal counterpart. The north part of the lake bottom declined gently to the centre of the lake, whereas the southern part dipped steeply. The maximum depth of the lake in 2005 was 17 m, which created a depression in front of the current ice margin. Moreover, buried ice still existed within some fragments of the lakeshore, islands, and at least to some extent at the bottom of the lake as well. Several small deltas have developed along the lakeshore (in places where the water was not deep), some of which were subsequently abandoned during glacier recession.

Evolution of the lateral moraines 1961–2013

Lateral moraines are the most prominent landforms in the Ragnarbreen foreland. They are built of significant amounts of ice covered by a relatively thin veneer (0.5–2 m thick) of debris. Since 1961, the lateral moraines were the most active landform assemblages within the glacier's foreland. Elevation lowering in the period 1961–2009 was at a maximum of –17 m (–0.35 m a^{-1}) compared to more than –90 m for the clean glacier surface. The volume decrease of lateral moraines was 4.3 million m^3. Most dynamic changes were related to debris flow activity and related dead-ice backwasting. Elevation decrease in areas without active mass movement processes was lower and interpreted as dead-ice downwasting.

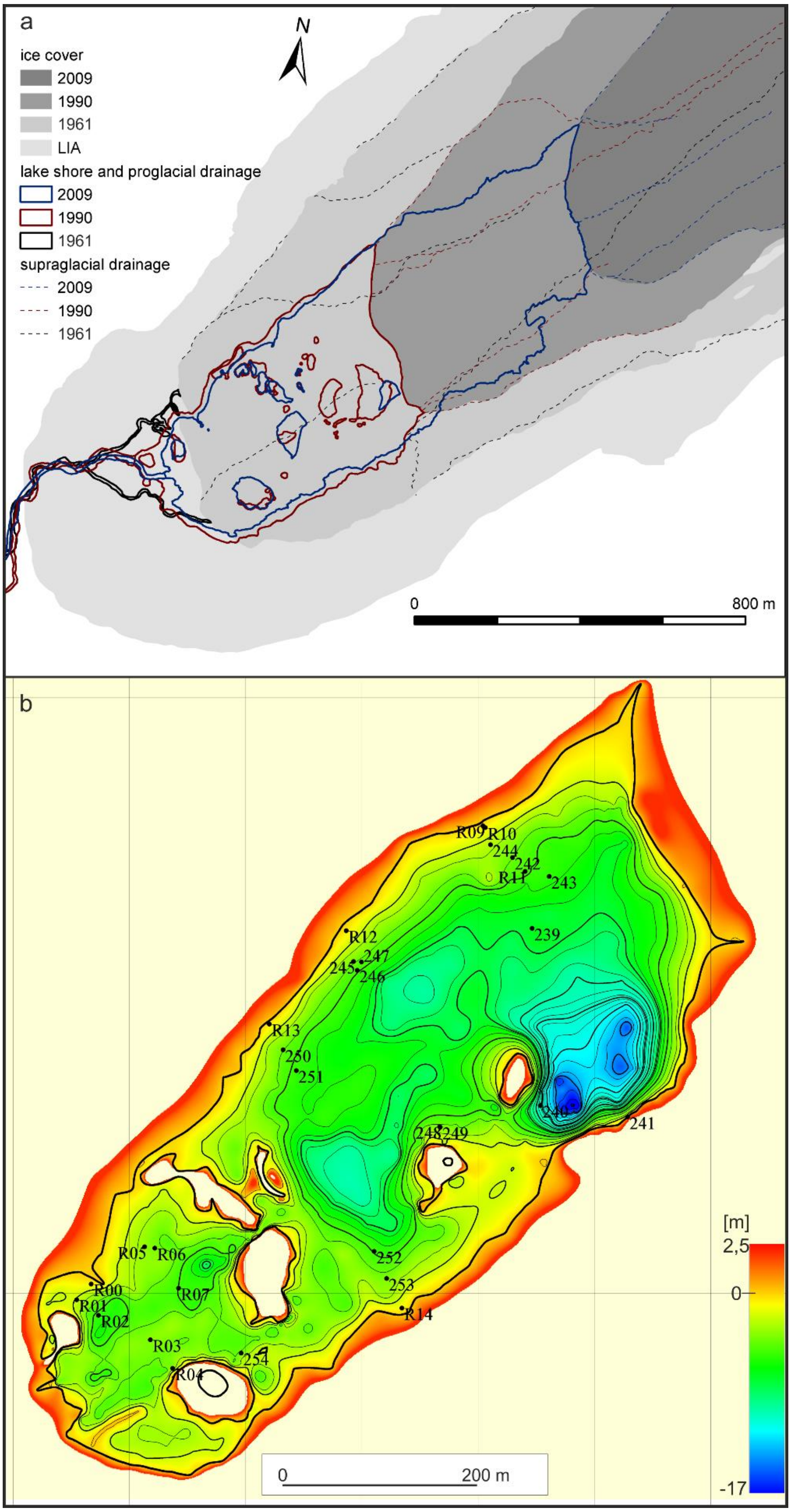

Figure 5. Evolution of the terminoglacial lake, Ragnarbreen, Svalbard; (a) lake shorelines and drainage for different dates; (b) lake bathymetry in 2005 (surveyed by J. Czerniawska, M. Ewertowski, W. Szczuciński)

In 2009, the lateral moraines extended along the glacier tongue up to 3000 m. The southern moraine was higher and wider than the northern one. Absolute height of the moraines varied from around 120 m to 350 m, whereas the relative height above the lake level or glacier surface was from 10 m (in the distal part of the moraine) to 100 m in (the place where the 2009 ice margin occurred). Upvalley, the relative height decreased, finally levelling with the glacier surface. The width of the southern moraine was between 210 m and 350 m. The northern moraine was between 10 m and 80 m in terms of relative height, and between 200 m and 300 m in terms of width. At present, debris delivered from the valley sides is stored in the depression between the lateral moraine and slope or within the lateral moraine. This situation effectively cuts off the supply of supraglacial debris to the glacier surface. Hence, there is no possibility to incorporate supraglacial debris into the glacier transport system and deliver it to the contemporary ice margin.

The lateral moraine's development is related to the presence and variability of distribution of the supraglacial debris cover. The occurrence of a sparse and thin debris layer on the ice surface can intensify ablation due to a reduction of albedo, but debris cover thicker than several centimetres can decrease or even stop the ablation of underlying ice by reducing insolation and radiation (Nicholson and Benn 2006; Kellerer-Pirklbauer *et al.* 2008; Nicholson and Benn 2013). The lateral moraine complex was divided into three parts according to its topographic characteristics in 2013. Hence, the outlines of each part have been located and superimposed on earlier aerial photographs in order to assess their temporal evolution since the earliest photographs dating to 1961. Each part has evolved through the same set of processes over time, so they can be regarded as stages in a quantifiable process-form continuum. Characteristics of the three parts of the lateral moraines and their changes 1961–2009 are as follows:

(A) Lateral moraine along the contemporary glacier surface (Figs. 6a, b, 7a). This part of the lateral moraines took a form of single crests adjoining the valley sides or, less often, separated from the valley side by a small depression, usually with a stream. The slope of the ridge was steep (>30%), but the profile was smooth. Only in some places were traces of water or debris flow observed. The transition from the slope of the lateral moraine (covered with debris) into the clean ice surface was distinct (usually a small supraglacial stream was flowing along these borders). Occasionally in some places, water flowing from the small patches of snow and ice located above the glacier went through the lateral moraine, thereby exposing the ice cores and showing how thin (usually less than 0.7 m) the debris cover was (Fig. 6).

(B) Lateral moraine along the lake (Fig. 7b). This part had a slightly different elevation profile, compared to the lateral moraine along the contemporary glacial surface. A single crest, 10–20 m wide, had developed adjoining the slope (in some parts, the crest was separated by a small depression). A steep scarp, 2-8 m high, had developed at the inner side. Exposed ice cores were often visible within these scarps. Below the scarp, slope of the lateral moraine declined steeply toward the lake level. Relief was highly rough with plenty of small hillocks, flow lobes, channels and headscarps (Fig. 6d, e). Dynamics of the geomorphic processes, mostly various types of debris flows, was much more intense in this part of the moraine. Observations during

the fieldwork showed that some fragments were remodelled during very short periods of time, from several hours to a couple of days.

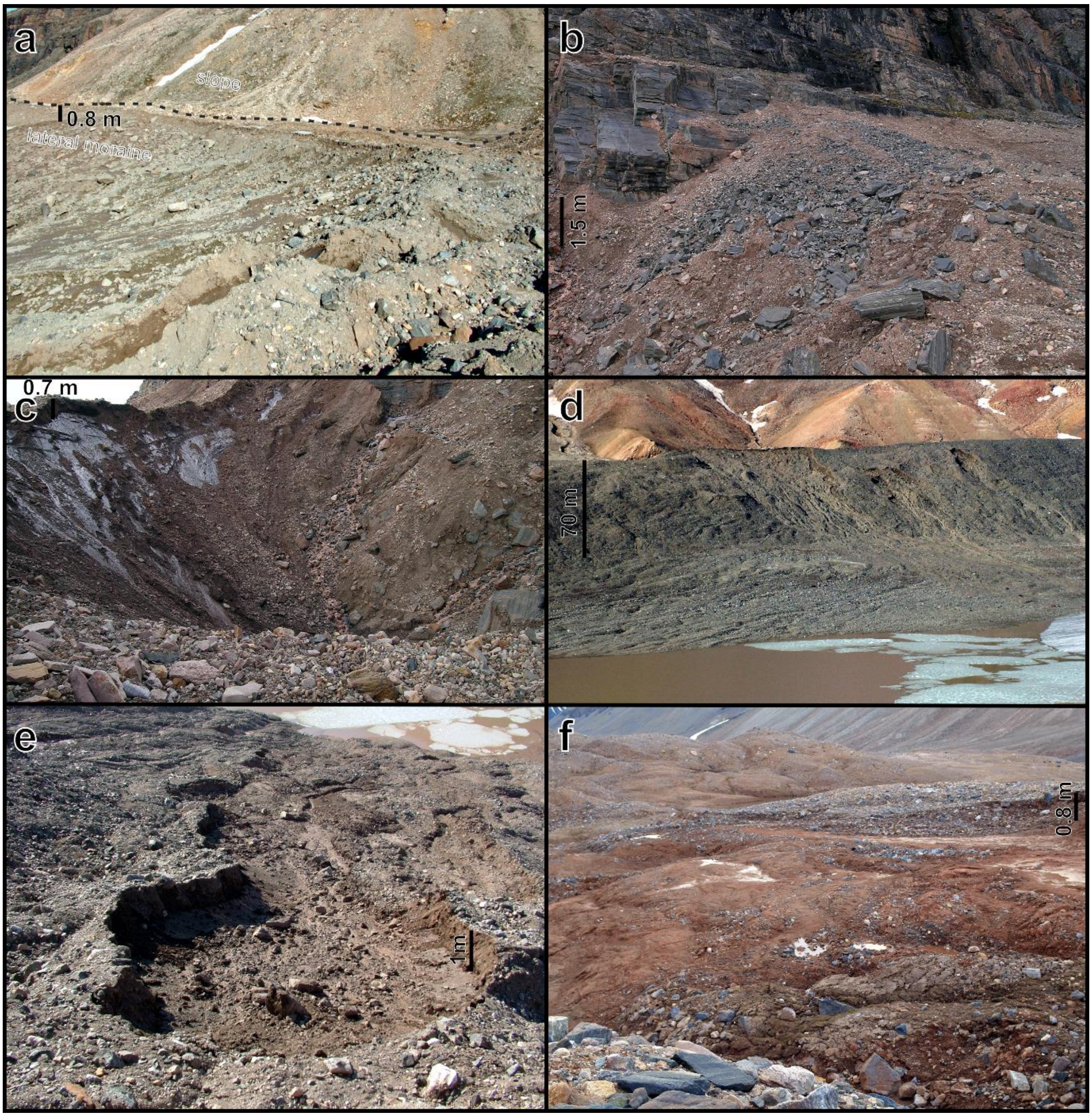

Figure 6. Lateral moraine of Ragnarbreen. (a), (b) gentle border between the valley side and lateral moraine; (c) stream going through the lateral moraine. Note how thin the debris cover is; (d) lateral moraine along the lake with visible debris flow activity; (e) fragment of the lateral moraine intensively remodelled by debris flows; (f) transitional zone between lateral moraine and end moraine complex; note that the relief is smooth and debris flow activity is much smaller

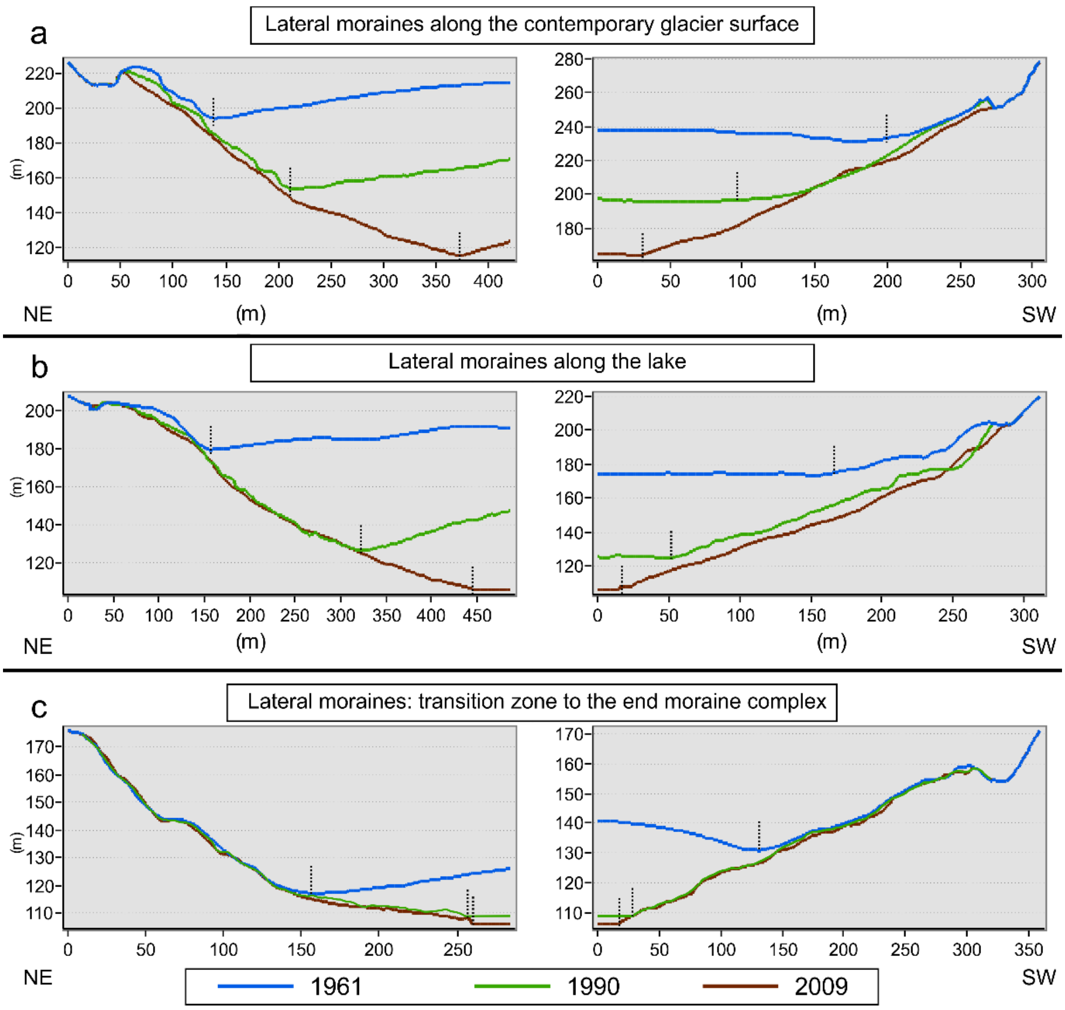

Figure 7. Elevation profiles across lateral moraines, Ragnarbreen, Svalbard. Vertical dotted lines mark location of the clean ice margin or the lakeshore

(C) Lateral moraine: transition zone to the end moraine complex (Fig. 7c). This part of the lateral moraine was the most stable. In terms of morphology, it consists of an outer, well-developed ridge, which was separated from the valley sides by a several-meters deep depression, filled with water that flowed under the lateral moraine as well as from the valley sides. The surface consisted of small hillocks, as well as mostly inactive lobes and channels' debris flows (Fig. 6f). Several small water ponds were also observed. In some fragments, debris flows were still active with the exposed ice cores remaining visible.

Five dominant lithofacies were recognized in the lateral moraines:

1) Massive boulders and gravels (Fig. 8a). This lithofacies was characterized by a large spread of clasts sizes from several centimetres to several meters in diameter. Individual clasts were angular or very angular. Empty spaces occurred between the larger clasts (sometimes filled with coarse gravels). Sandy or clay matrix was absent. This lithofacies was found mostly in the lateral moraine along the contemporary glacier surface.

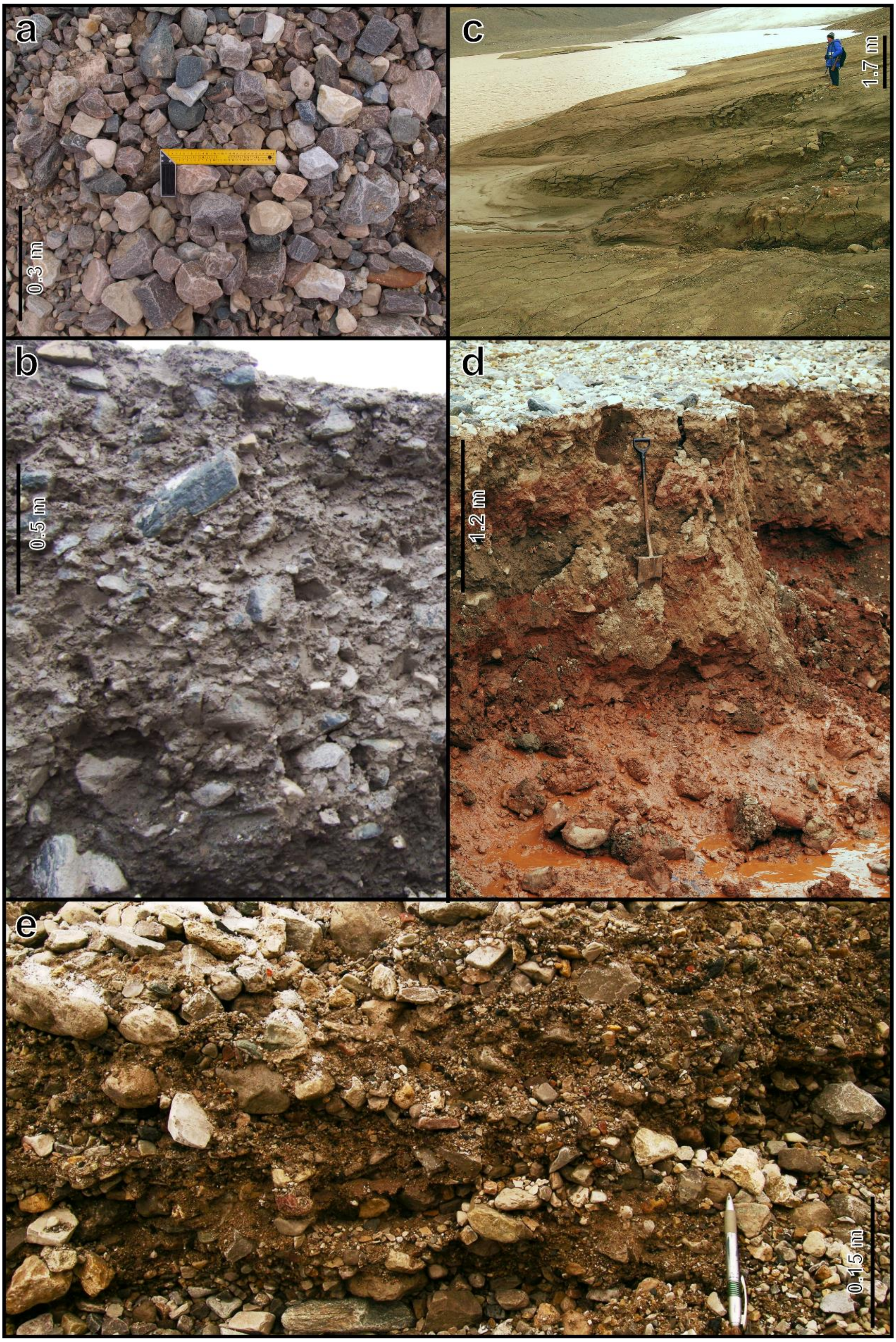

Figure 8. Examples of lithofacies, Ragnarbreen foreland, Svalbard: (a) Massive boulders and gravels, (b) Massive, clast-supported diamicton (photo: I. Szuman), (c) Massive, diamictic sand, (d) Massive, matrix supported diamicton with varied clast content, (d) Massive, diamictic gravel

2) Massive, clast-supported diamicton (Fig. 8b). This lithofacies comprised a very high proportion of coarse gravels (average diameter 10 to 25 cm, sometimes exceeding 50 cm). Clasts were angular, but were better rounded than in lithofacies of massive boulders. Fine- or medium-grained (mostly sandy) matrix filled up the space between individual clasts. It was the most common lithofacies in the lateral moraines. Massive, clast-supported diamicton built most of the ridges occurring between the debris flows and sometimes was also found in the debris flows themselves.
3) Massive, diamictic sand (Fig. 8c). Sands of varied grain sizes dominate this lithofacies. The matrix was fine, clayey-silty and lower in proportion than in lithofacies of massive, matrix supported diamicton. Individual clasts of fine-and medium gravel were very rare. Massive, diamictic sand usually constituted lobe-shaped deposits or small fans. Quite often, individual lithofacies of diamictic sand were superimposed on each other.
4) Massive, matrix supported diamicton with varied clast content (Fig. 8d). Three subtypes were observed depending on the amount of the clast content (low, moderate, high). The subtype with a moderate amount of clasts was the most commonly observed. The matrix comprised mainly of sands and fines. Clasts, usually subangular, were scattered within the matrix. Massive, medium-grained, matrix supported diamicton was the most common in that part of the lateral moraine which was the transition zone to the end moraine complex.
5) Massive, daimictic gravel (Fig. 8e). Clasts in this lithofacies were usually from 5 to 15 cm in diameter and very angular or angular. Coarse-grained matrix occurred between clasts. Matrix was usually sandy or sandy-gravelly, but the amount of matrix was much less than in lithofacies of massive, clast-supported diamicton. Massive, diamictic gravel was usually found in the area of debris flow lobes and near the debris flow channels.

Elevation changes within the end moraine complex (1961–2009)

The end moraine complex consisted of arcuate chains of hillocks and ponds (Figure 9). The complex was elevated up to 35 m above the sandur level and 20 m above the lake level. Individual hillocks reached usually 10–50 m in length and 2–4 m in height. Some of them had characteristic sharp-crested shapes. Many lobes of old debris flow deposits were observed; however active debris flows were rare and restricted to the areas undercut by streams.

The depositional environment of the end moraine complex has been described in detail previously (Ewertowski *et al.* 2012). Hence, only elevation changes are discussed here. Despite being ice-cored, the end moraine complex has been relatively stable over the last 50 years (Fig. 10). Maximum elevation changes did not exceed 10 m in the period 1961–1990 and 11 m in the period 1990–2009; however such big changes were restricted only to small areas. Major transformations were observed only in places with high stream activity and rare dead-ice exposures. Most of the end moraine complex showed no changes or even slight accumulation in the period 1990–2009. Overall decrease of volume of the end moraine complex in the period 1961–2009 was 0.4 million m^3.

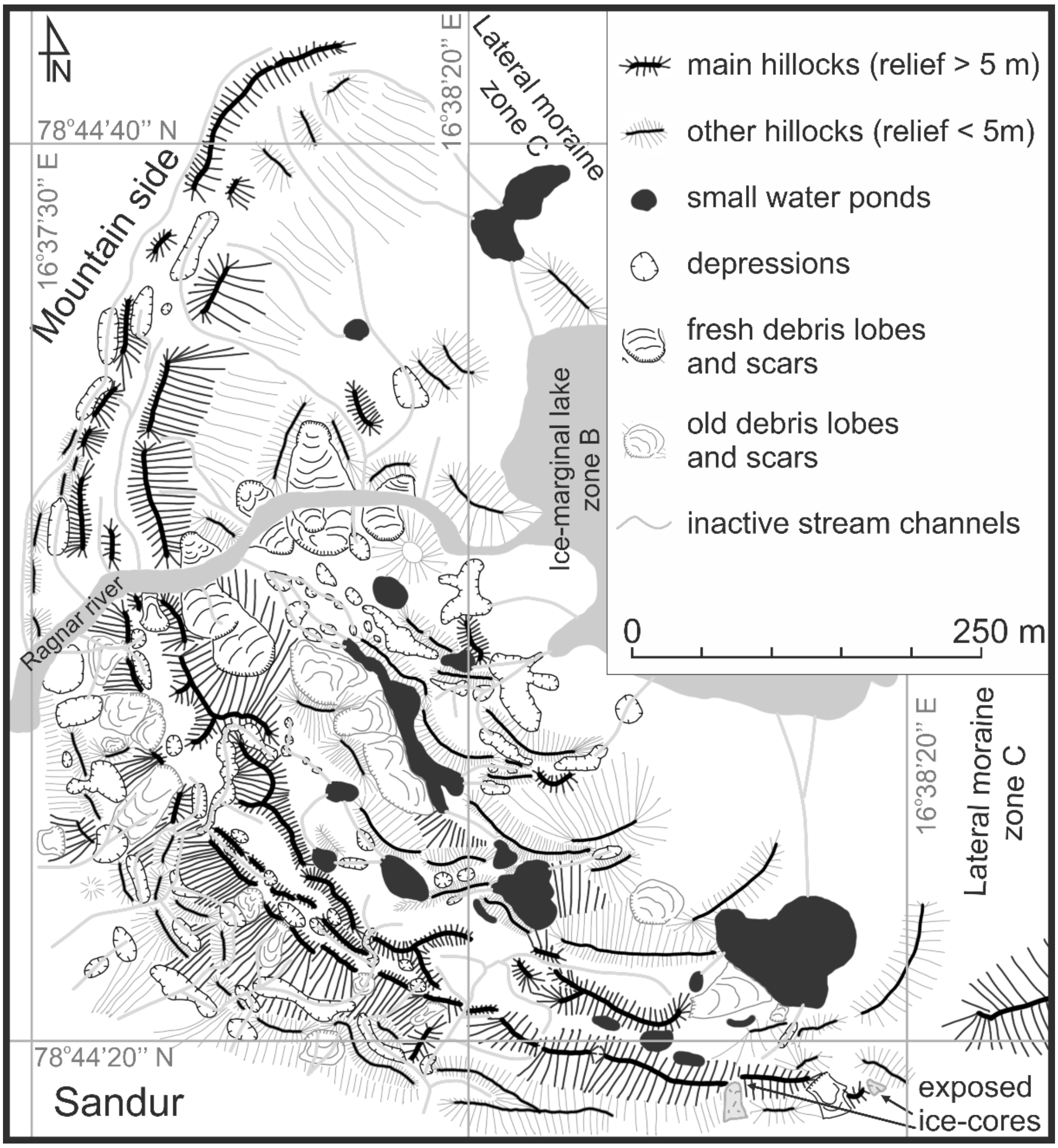

Figure 9. Morphological sketch of end moraine complex showing arcuate chains of ridges and depressions, Ragnarbreen, Svalbard; reproduced from: Ewertowski et al. (2012), Zeitschrift für Geomorphologie, 56(1):53-74 with permission from E. Schweizerbart'sche Verlagsbuchhandlung, www.schweizerbart.de

Summary of dynamics of proglacial landforms

Based on evidence from multi-temporal DEMs and orthophotos, the clean ice surface (i.e. part of the glacier snout with no debris cover) was the most dynamic, with mean elevation changes of –1.14 m a^{-1} during 1961–1990 and –1.23 m a^{-1} during 1990–2009 (Table 4). Less active were lateral moraines. These experienced mean elevation changes of –0.15 m a^{-1} in 1961–1990 comparing to –0.08 m a^{-1} in 1990–2009. The most stable landform was the end moraine complex with annual lowering rates of less than 0.05 m a^{-1}.

Data provide in Table 4 suggest that in some part of the moraines elevation increased (with maximum values up to 5.7 m). Such large increases of elevation should be treated with cautions and are most probably related to local DEM errors (caused for example by shadows on aerial photographs). However, localized positive changes in elevation were possible and were also observed during fieldwork. For example, transfer of debris by mass wasting processes can cause decrease of elevation in areas of high topography and consequently increased elevation in local depositional centres.

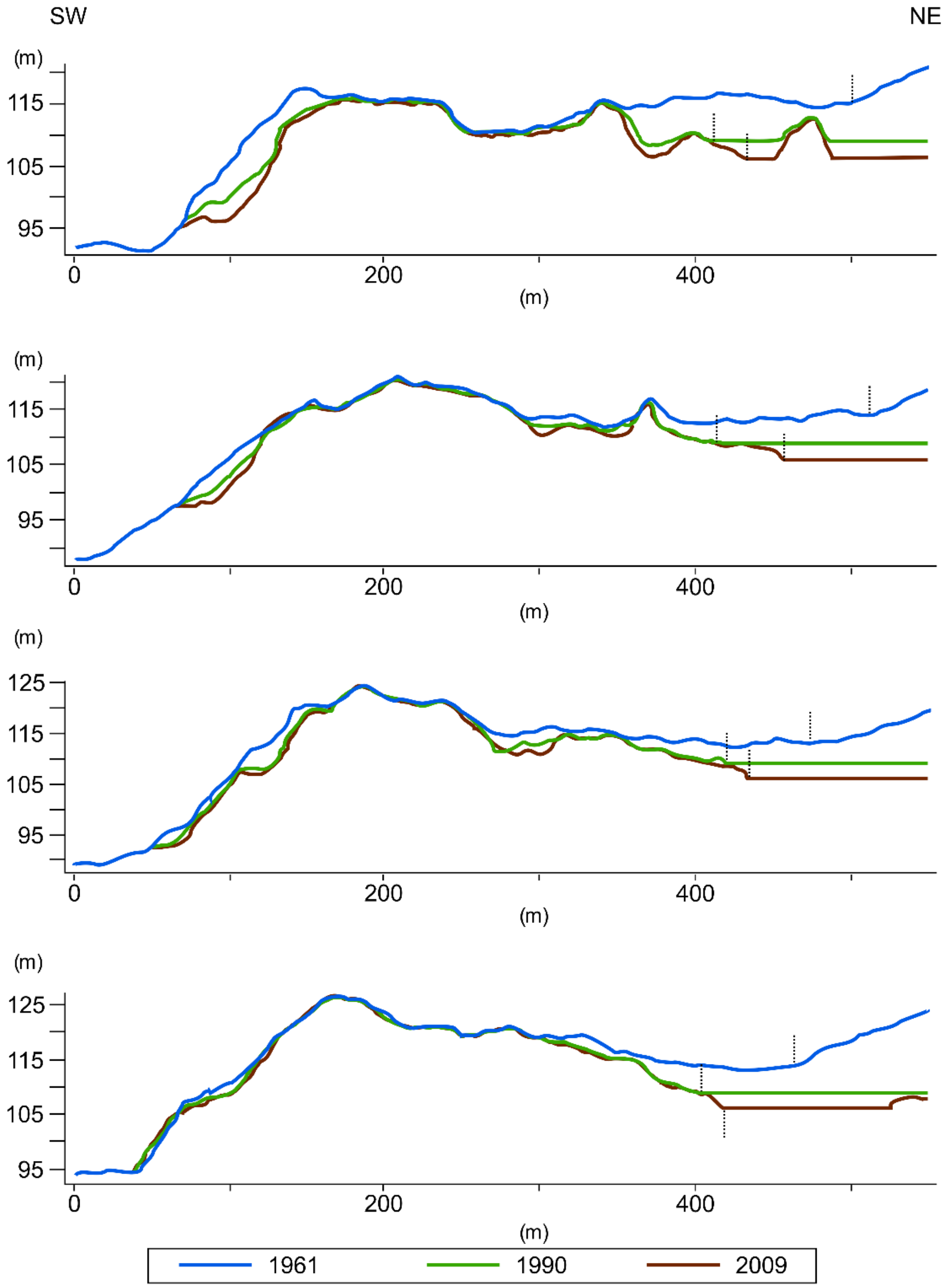

Figure 10. Elevation profiles across the end moraine complex, Ragnarbreen, Svalbard. Vertical dotted lines mark location of the clean ice margin or the lakeshore

Discussion

Evolution of Ragnarbreen proglacial landforms 1900(1920)–2013

Recent (1900/1920 to 2013) transformation of the Ragnarbreen landsystem shows the large role played by local topographic conditions in glacial landform creation. Three main stages in the snout and landform transformation are proposed and interpreted based on changes in geometry and geomorphology (Fig. 11).

1900/1920–1961. As there are no remote sensing data concerning this period, information is only hypothesised based on field observation and geomorphological relationships between moraines and valley sides. Following the Little Ice Age (LIA) air temperature in Svalbard increased (Nordli *et al.* 1996; Førland and Hanssen-Bauer 2003) and the equilibrium line moved upwards, leading to glacier negative mass balance. Location of the ice surface in 1961 comparing to the height of the lateral moraine suggests that at the beginning of deglaciation after the termination of LIA, the loss of ice mass was mainly due to glacier thinning. During this period, the glacier retreated from its maximum extent, exposing the end moraine complex and started to develop lateral moraines in the frontal part of the glacier.

Because the debris-covered lateral moraines melted to a lesser degree than the relatively clean ice surface, the glacier profile probably changed from convex to concave (based on current relation of high elevated lateral moraine crests and low lying clean ice surface, Fig. 6). The genesis of the end moraine complex proposed by Ewertowski et al. (2012) highlighted the role of topographic inversion in this landform assemblage. However, the new elevation data created for this study suggested that no significant topographic inversion occurred since 1961, suggesting that other processes were also involved in the end moraine complex's construction and degradation. Additionally, due to the negative mass balance of the glacier, its active part flowed in a tensional mode, such that the thrusting of debris-rich ice layers in the frontal part of the glacier ceased. The new data can also suggest some role for thrusting in the genesis of the end moraine complex, which can be supported by the occurrence of arcuate, sharp-crest ridges within the moraine, and the depression now occupied by the lake, which can give an impression of a hill-hole pair. Progressive melting of ice cores in the end moraine led to reduction of moraine height and formation of small water ponds.

1961–1990. In this period, the end moraine complex was relatively stable. The most important aspect of landscape transformation was the development of the terminoglacial lake, which also covered part of the glacier snout. The blockage of drainage, creation and expansion of the lake probably enhanced the glacier retreat rates. Moreover, along with glacier recession, the terminoglacial lake acted as a sedimentary trap, effectively reducing sediment delivery to the end moraine complex. However, sediments were and still are transported through the moraine complex by the Ragnar stream, which developed an outwash fan and also contributed to sediment delivery to the tidal flat in the Petuniabukta.

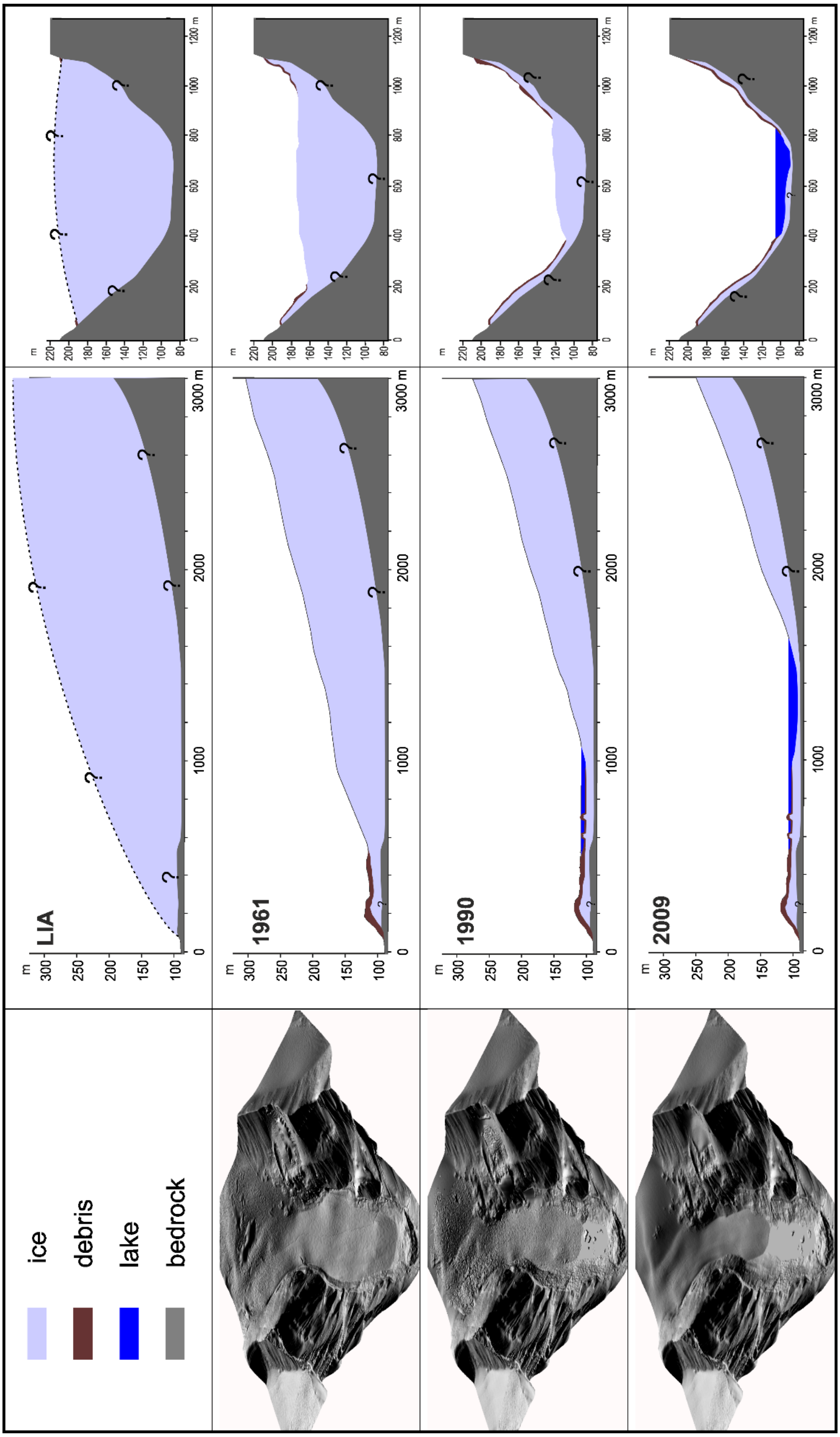

Figure 11. Transformations of Ragnarbreen landsystem (1961–2013). Surface profiles based on DEM. Bedrock topography, thickness of the debris covers and LIA ice profile are, however, hypothetical

A further lowering of the clean ice surface also had an important role in lateral moraine formation. Lowering of the ice surface, which was most significant in the area of the contemporary lake (Fig. 3), accelerated elevation changes between the lateral moraines' tops (covered with debris) and the clean ice surface. This led to the creation of unstable slopes and intensification of mass wasting processes.

1990–2013. The terminoglacial lake constantly expanded in area. Moreover, the lake filled up the deeper part of the depression, which meant that the overall volume of the water storage within the lake significantly increased. A further lowering of the glacier surface, linked with changes in the shape of its transverse elevation profiles, caused further intensification of mass wasting of the lateral, ice-cored moraines. The deposits have been sorted by meltwater and subsequently mixed by the mass flows, often over a very short distance.

It is typical that deposits of the moraine complexes have been repeatedly transformed. As suggested by other studies (Bennett *et al.* 2000; Lukas *et al.* 2005; Schomacker and Kjær 2008), debris flows are among the most significant mechanisms of reworking and redeposition in the glacier forelands on Svalbard. A similar situation can be seen in Ragnarbreen. However, although the end moraine complex of the Ragnar glacier is still ice-cored, fresh debris flows were not ubiquitous during last ten years. After active ice retreat, other processes such as aeolian activity and lacustrine and fluvioglacial sedimentation may be important in the displacement of glacigenic sediments within the end moraine complex (cf. Ewertowski et al., 2012). Some lowering of the end moraine complex (Fig. 5) is therefore related to the downwasting of dead-ice rather than backwasting. A similar situation was observed on the outer moraines of Austre Lovénbreen (Midgley *et al.* 2013). At the same time, active debris on lateral moraines suggest a spatio-temporal shifting of slope instability zones over time.

Development of the lateral moraines

The transformation of the ice-cored lateral moraine of Ragnabreen in the period 1920–2013 can be divided into four phases:

(1) Lateral moraines developed probably when glacier reached its maximum extent during the LIA. During the maximum extent of ice cover, the transverse profile of the glacier was most likely convex. Weathering contributed to the large quantity of debris from the valley sides, which was delivered to the glacier surface. The debris accumulated on the lateral margin of the ice surface, creating a continuous but not very thick cover of angular supraglacial debris. Climate changes after the end of the LIA (1900/1920) led to differentiated ice melting (debris-covered ice versus clean ice surface), causing a change to a concave glacier transverse profile and the creation of ice-cored lateral moraines, which was higher than the clean ice surface. Moreover, streams appeared between the lateral moraines and valley sides and separated the lateral moraines from the debris source.

(2) A lowering of the clean ice surface caused slope instability (Fig. 12a). This led to the development of sporadic debris flows and the exposure of the ice cores. However, debris cover was still thick enough to stop the majority of dead-ice from melting (i.e. the debris cover's thickness was greater than that of the permafrost active layer).

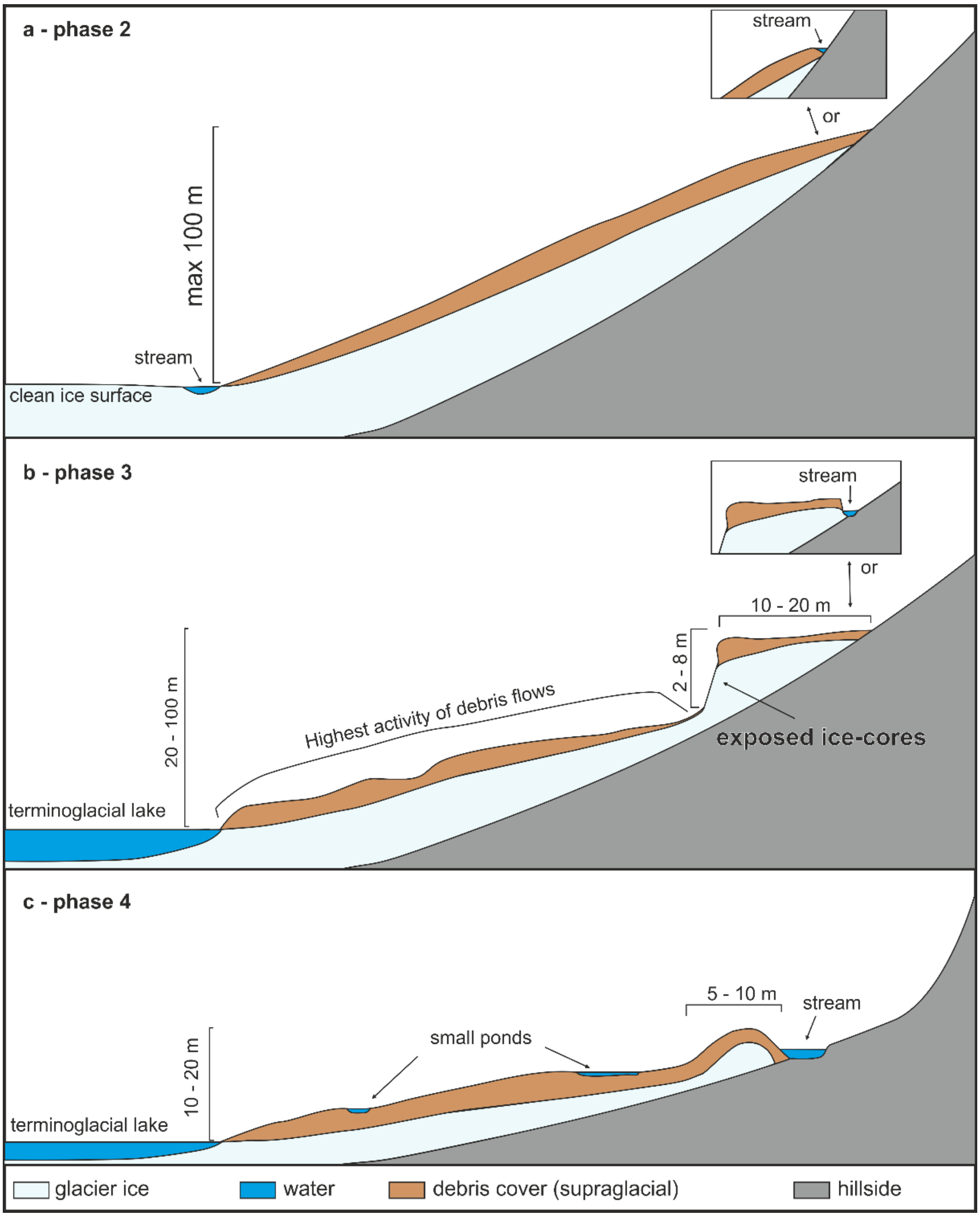

Figure 12. A model of the lateral moraine development, Ragnarbreen, Svalbard; (a) phase 2; (b) phase 3; (c) phase 4. Further explanation in the text

(3) After reaching a certain threshold in elevation difference between the clean ice surface and debris-covered lateral moraine, slopes became steep enough to initiate common mass movements (Fig. 12b). These processes form some sort of coupled, cyclic continuum: The development of debris flows exposed the ice cores, which increased the amount of meltwater and in turn the activity of debris flows. The initial mobilization of sediments in a specific place can be caused by: (1) saturation of debris with water due to precipitation; (2) saturation of debris by water melted from dead-ice; or (3) increasing slope steepness due to previous debris flow or undercutting by rivers (Lawson 1979; Krüger and Kjær 2000; Kjær and Krüger 2001; Schomacker 2008; Schomacker and Kjær 2008).

(4) Debris re-sedimentation during the later phases led to substantial levelling of the elevation profile and reduction in elevation differences (Fig. 12c). Relatively thick debris cover (up to 2 m) and a lack of steep terrain sections caused lowering of geomorphic activity, while dead-ice melting clearly decreased compared to the previous stages. On gentle slopes, the debris cover was thick enough to stop or at least significantly slow down the melting of ice cores. Thus, active debris flows were rare, limited only to the areas where ice cored exposures were caused by undercutting by streams.

Conclusions

Field-based observations (2005–2013), combined with DEMs, orthophoto creation and interpretation, allowed for the quantification of recent glacial landform transformations in the proglacial area of Ragnarbreen. Assessment of landforms and glacier snout changes indicate that in the studied high-Arctic setting the glacier showed a 135 million m^3 decrease in volume during the period 1961–2009. Despite the fact that most of the landforms are ice-cored, the transformation of landforms was at a much lower magnitude, reaching 5 million m^3 of volume loss (1961–2009). Glacier retreat rate was higher after 1961 than before it, and since the end of the LIA the ice cover has decreased by 26%.

The resultant landsystem at Ragnarbreen comprises three main components: lateral moraines, end moraine complex, and terminoglacial (partly supraglacial) lake. In terms of landscape alteration, the most important event was the creation of a terminoglacial lake, which acted as a sedimentary trap and at the same time probably accelerated glacier mass loss. The second most active component was the lateral moraines.

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